Water balance and contributing area of a Pocosin watershed in North Carolina, USA.

Thesis submitted for a M.Sc. degree in Forest Sciences and Business

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June 2014

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Työn nimi — Arbetets titel — Title
Water balance and contributing area of a Pocosin watershed in North Carolina, USA

Oppiaine —Läroämne — Subject
Forest ecology and management

Työn laji — Arbetets art — Level	Aika — Datum — Month and year	Sivumäärä — Sidoantal — Number of pages
MSc thesis	May 2014	49

Tiivistelmä — Referat — Abstract

The pocosins of the North Carolina Atlantic Coastal Plain (ACP) region play a vital role in controlling hydrologic patterns and determine wetland development and ecosystem structure and function. However, these low-lying, shallow water table forested watersheds have received little research attention due to the relative scarcity of long-term monitoring data, watershed delineation challenges and scarcity of unmanaged sites. Without holding anthropogenic activities constant, specific hydrological processes such as runoff generation, are difficult to describe because of drainage network influence on soil water storage.

The principal goal of this study was to develop and test a water balance framework for investigating runoff characteristics and the active watershed area contributing to runoff of a managed pocosin watershed in the ACP. Using a water balance approach, the first objective was to calculate monthly and annual water balances with particular emphasis on deriving, mathematically, the area contributing to runoff. The second objective was to explore the concept of variable contributing area by converting discharge measured at the outlet to runoff using a watershed area. The watershed area needed to produce runoff values that result in closure of the water balance equation represents the variable runoff contributing area. The calculation was done for different temporal periods: monthly, seasonal, and annual and runoff contributing area values compared to the topographically defined watershed area.

The results of the study indicated that water balance components were generally in good agreement and closure tended to occur at longer time scales, decreasing for shorter periods. Lack of system closure at shorter temporal scales suggested that the contributing area to runoff varied and differed from the topographically defined watershed area. Active contributing area clearly varies temporally but on average is estimated to be approximately 600 hectares. Regression predicted watershed size was smaller than expected which could have been due to the difference between measured and predicted streamflow. The extent at which the active contributing area fluctuates depends on the compounding uncertainty of water balance components.

Avainsanat — Nyckelord — Keywords
Water balance, watershed management, contributing area, pocosin

Säilytyspaikka — Förvaringsställe — Where deposited
Viikki Science Library, University of Helsinki e-thesis

Muita tietoja — Övriga uppgifter — Further information

ACKNOWLEDGMENTS

I would like to express my appreciation to my advisory committee composed of professors at the University of Helsinki and Atlantis Programme partner North Carolina State University: Dr. Michael Starr, Dr. Ryan Emanuel, Dr. Glenn Catts, and Dr. Joe Roise. Thanks for giving me the opportunity to be creative with a comprehensive dataset for Hofmann Forest. Special thanks to Dr. Michael Starr and Dr. Ryan Emanuel for their time, patience, and understanding through the long and arduous process of creating a study. Dr. Glenn Catts provided invaluable expertise on Hofmann Forest and a broad spectrum of knowledge on the range of topics my thesis addresses. The kind folks at Hofmann Forest, especially Joe Jarman, made my project possible by collecting (and continuing to) an excellent operational and hydrologic dataset. Hofmann Forest personnel were not aware that someone would come along and create a study from their daily work.

The most special thanks I owe to my best friend, partner, and now my wife. Ashley, who gave me your unconditional support and love through this long process.

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1. INTRODUCTION

The primary goal of this study was to investigate the water balance and the active watershed area contributing to runoff of a managed pocosin watershed in the Atlantic Coastal Plain of North Carolina, U.SA. The forest hydrology of pocosin watersheds have received little research attention compared to upland watersheds due to the relative scarcity of long-term monitoring data, relative absence of unaltered sites, and watershed delineation challenges. Without holding anthropogenic activities constant, specific hydrological processes such as runoff generation, are difficult to describe because of the managed drainage system influence on soil water storage. In this study, specific hydrological processes, including storage and runoff relationships, as well as the active contributing area of a pocosin watershed were evaluated using a water balance framework. Several statistical approaches are presented to investigate the closure of the water balance and the varying active contributing area.

1.1. Pocosins

Pocosins, literally meaning “swamp on a hill,” and adopted from an Algoquian Indian term ‘Poquosin’ (Tooker 1899) are endemic landforms to the southeastern Atlantic Coastal Plain (ACP) of the U.S.A. (Richardson, 1991). In the wetland classification system, pocosins are generally classified as palustrine (inland, freshwater swamp) ecosystems with either scrub-shrub or forested vegetation (Richardson, 1991). Pocosins are a unique type of seasonal forested wetland, characterized by a distinctive hydrology of long hydroperiods, temporary surface inundation, periodic burning of fire-adapted ericaceous vegetation, and highly organic soils of sandy humus, muck or peat (Wells, 1928; Woodwell, 1958; Kologiski, 1977; Skaggs et. al., 1991). They occur at drainage heads on broad, flat divides and have low topographic relief with relatively high water tables which fluctuate substantially throughout the year (Richardson, 1983).The deep, moist soils are the result of centuries of organic matter buildup under anaerobic conditions (Richardson, 1991). Pocosin sites can be remarkably fertile if drained and are well suited for crop production and forestry. Pocosins and associated wetland forests nearby are among the most productive in the US in terms of softwood production (Allen and Campbell, 1988; Campbell and Hughes, 1991). To realize their site productivity potential, water management, site preparation, and infrastructure improvements are essential for timber production on these sites (Askew and Williams, 1979). The results of water control activities, specifically ditch network construction, generate pronounced hydrologic responses in the form of sustained streamflow generation and reduced soil water storage, which are the result of dynamic interactions between soil, climatic conditions, and vegetation (Skaggs et al., 1991 and Sun et al., 2008). As a result, most pocosins contain vast ditched, drainage networks and few pristine pocosin systems remain (Richardson, 2003). Some have suggested that the increased ditch networks facilitate ecological connectivity between terrestrial and aquatic ecosystems from the lateral transport of water and nutrients (Pringle, 2003). Coastal estuarine habitats are hydrologically connected to the pocosins upstream (Richardson, 1991), and pocosins serve to moderate outflow frequency and magnitude. Therefore the drainage and forest management of pocosins have the potential to affect downstream wetland functions. However, the negative effects of forestry activities on water quality in the southeastern U.S.A. are generally less than the effects of other activities such as agriculture and urbanization (Dissmeyer, 2000).

Few studies have examined the geographic distribution of pocosins, but historically they are thought to have once covered more than 400,000 hectares from Virginia to north Florida (Richardson, 2003). Seventy percent of the pocosins in the US occur in North Carolina and made up more than fifty percent of the North Carolina freshwater wetlands in recent decades (Richardson, 1983). The exact distribution of pocosin management (i.e. area drained, forest plantation conversion) has not been mapped recently, but the relative absence of large areas of native pocosin vegetation suggest active management including new ditch construction and regular ditch maintenance of pocosins is widespread.

1.2. Watershed characteristics

Pocosin forests of the ACP are defined by a seasonally high water table, low topographic relief and relatively fertile soils (Richardson and McCarthy, 1994). Due to their position on the landscape, pocosins serve as drainage heads and do not receive drainage waters from upstream (Richardson, 1983, Shepard, 1994). The low topographic relief characteristic of pocosins means that hydraulic gradients are small and flow is strongly influenced by water table depth and storage (Harder et al., 2007). The high water table and seasonally wet site conditions on pristine pocosins can limit agricultural and forestry production and therefore most sites are drained (Figure 1) (Skaggs et al., 1994). The ditch network lowers the water table and keeps it at a more constant level (Amatya et. al., 1996), stabilizes potential nutrient export downstream (Amatya et. al., 1998), and allows for equipment operability (Campbell and Hughes, 1991). The ditch network effectively functions as first-order streams regulating watershed outflow (McCarthy et. al., 1991; Amatya, 2003; Skaggs et al., 1991). A typical ditch network on a drained pocosin may extend beyond the topographically delineated watershed boundaries (Figures 3 & 4).

Figure 1: A recently constructed ditch on Hofmann Forest in the Southeastern Coastal Plain of North Carolina, U.S.A. (Source: Eric Jaeschke).

The ditch network on a pocosin site effectively steepens the water table and promotes the lateral movement of shallow groundwater towards the ditch. Shallow subsurface flow depends on the hydraulic gradient as described by Darcy's Law (Dingman, 2002). The hydraulic gradient helps quantify the rate of groundwater flow by determining the energy balance shift from high to low (Hornberger, 1998). An increase in hydraulic gradient caused by the ditch network has the potential to pull in water from larger or smaller areas depending on the density of the ditch network. Shallow groundwater flow in pocosins could also be determined by macropores and cavities in the upper soil layers, thus enhancing permeability in the near-surface soil and allowing for high rainfall retention. Organic histosol soils are able to hold a large amount of water relative to the oven-dry mass of the soil, 300-3000% by some estimates (Verry et al., 2011). Organic soils in the lower layers experience low permeability due to the predominantly saturated conditions and perched position above a confining layer of clay and sand. Deep percolation or groundwater discharge to the regional aquifer is considered negligible (Richardson, 1983). Because of the low relief there is a tendency for surface water to accumulate until evaporated or transpired by vegetation (Richardson and McCarthy, 1994) and overland flow seldom occurs on these sites (Campbell and Hughes, 1991). The permeability characteristics and high water table of pocosin sites results in their characteristic "flashy" runoff response to rainfall (Skaggs et al., 1991). This phenomenon is illustrated by a steep flow duration curve relative to non-pocosin watersheds (Figure 2). On a pocosin site the ditch network regulates discharge in part by roughness of the ditch surface and can be limited by the capacity of man-made outlets (Amatya et al., 1996).

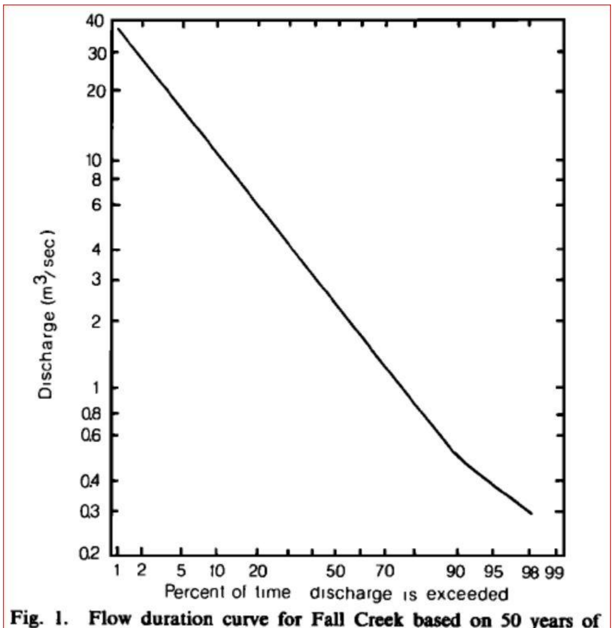


Fig. 1. Flow duration curve for Fall Creek based on 50 years of data (U.S. Geological Survey, Ithaca, New York).

Figure 2: Generic flow duration curve (*USGS*, Ithaca, NY, 1979 as cited in Johnson, 1979)

Ground water levels and soil moisture in managed pocosins is a balance between precipitation inputs and evapotranspiration outputs, with runoff playing a minor role

(Richardson and McCarthy, 1994). Several studies in the ACP region have found runoff in the ditch network to range from 10-34% of precipitation (McCarthy et al., 1991; (Richardson and McCarthy, 1994). Forest harvesting allows the water table to rise and soil scarification transforms soil moisture conditions until vegetation reestablishment, especially in the absence of an actively draining ditch network (Lebo and Herrmann, 1998; Amatya et. al., 1996). The harvest activities or conversion of natural vegetation to plantation also temporarily increases stream flow, reduces ET, and interception and infiltration rates (Richardson and McCarthy, 1994; Lebo and Herrmann, 1998; Dissmeyer, 2000). The hydrological recovery or return to baseline conditions after forest management activities depends on the ability of vegetation to re-initiate normal evapotranspiration functions (Blinn and Aust, 2004). Evapotranspiration is high on pocosin sites, consuming 50-90% of the incident precipitation (McCarthy et al., 1991; Amatya et al., 1997; Sun et al., 2002). Actual rates of ET on pocosins vary seasonally with temperature, but also depend on leaf area index (LAI), forest canopy structure, stand density, and root zone soil water content (Richardson and McCarthy, 1994). Because of high water tables, the rooting zone on pocosin sites is also close to the surface, with water available to plants during most of the year (Amatya et al., 1996). Because water availability is usually not limiting, actual evapotranspiration can be assumed to equal potential evapotranspiration for pocosin sites.

1.3. Watershed contributing area

The contributing area of a watershed may generate runoff following complex interactions between topography, soil, climate and vegetation (Dunne and Black, 1970; Dunne, 1978). The runoff response of a watershed to precipitation events is a function of soil saturation levels, evapotranspiration demand of the vegetation and topographic gradients (Nippgen et al., 2011). Previous studies of drained pocosin hydrology (ACP) have established that the ditch drainage network can be a major driver of runoff behavior (Amatya et al., 1996; Amatya et al., 1997). The ability of the ditches to direct water off the landscape is controlled part by the density of the ditch network as well as the hydraulic gradients caused by ditch depth (Amatya et al., 1997). In this sense, the ditch network of a managed pocosin site functions to connect different parts of the landscape.

If the ditch network is considered a first order stream network, then the size of the area it drains should be close in size with the topographical watershed size. Phillips et al. (2011) suggest a differentiation between the gross drainage area and active contributing area, as the active stream network rarely is synonymous with the drainage network of all streams within the topographically delineated watershed. Therefore, saturated active areas that can generate runoff are not necessarily active contributing areas that are connected to the watershed outlet (Ambroise, 2004). The variable source area (VSA) concept offers a framework to explain some the mechanisms of streamflow generation at the watershed scale (Hewlett and Hibbert, 1963). However its usefulness for a site on the ACP is limited due to initial validation at a southern Appalachians mountain site and a bias towards importance of near-stream saturated zones (McDonnell, 2003). A more recent study suggested that streamflow generation (as measured by select storm events) and the variable source area were influenced by the extent of soil saturation (Sun et al., 2009).

The previous influential work in forest hydrology provides a basis to build a theoretical framework explaining runoff generation and a contributing source area concept in pocosins. Low topographic relief indicates that all parts of the drainage system would need to be saturated and connected to other saturated pathways in order to contribute runoff to the outlet. Seasonal soil moisture patterns that define the shallow flow processes (directing water to the ditch network) may trigger connectivity. Periodic, larger than normal rainfall events or sustained high-intensity events may push the watershed over the connectivity threshold. When the threshold is reached, the watershed is forced into a fully connected state (ditches are fully engaged), potentially drawing water from outside the topographically defined watershed area (W_D). In contrast, during dry periods the connections have the potential to disconnect due to the flat topography, high soil water absorptive capacity and evapotranspiration demands. Regardless of the extent of ditch network engagement, the adjusted (steepened) hydraulic gradient likely pulls in shallow groundwater from outside W_D, thus promoting a fluctuating contribution area. The largest ditches at Hofmann Forest are generally adjacent to the road network that is well connected to the drainage network directly and indirectly.

Shallow gradient changes of pocosin sites usually force ditch network flows in one direction, connecting various extensions that cease at the watershed boundary. After high intensity and duration rainfall events (i.e. tropical storms), Hofmann Forest personnel have observed water within the ditch at the perceived watershed boundary flow in the opposite direction away from the watershed outlet. This area of ditch flow reversal likely has a certain threshold when flows from outside the topographically delineated boundary (W_D) interact with flows inside, suggesting a variable watershed boundary. If the watershed boundary is fluctuating due to ditch network engagement, then the active contributing area to the watershed outlet is also changing. A water balance approach can be used to approximate the difference between W_D and the fluctuating active contributing area.

Major observed or estimated water fluxes such as precipitation, evapotranspiration, and soil water storage changes through space and time generate the variations in runoff response. The relative difference between soil water storage and runoff rate gives clues about the responsiveness of a watershed to precipitation events (Spence, 2007), and this relationship is intimately related to the dynamic drainage network. The change in soil water storage controls when runoff occurs and therefore storage influences runoff efficiency (Wooding, 1965). Since runoff is intimately related to storage, it can be speculated that precipitation indirectly causes expansion or shrinking of the ditch network (Moussa et al., 2002) because storage has an influence on overall network connectivity (Bracken and Croke 2007).

1.4. Watershed Balance

The water balance equation is an account of the inputs, outputs and changes in soil water storage for an area, typically a watershed, over a particular time period and expressed as depths of water (Dingman, 2002). The water balance equation is as follows:

$$\mathrm{P} = \mathrm{ET} + \mathrm{R} + \Delta\mathrm{S} \tag{1}$$

where P is precipitation, ET is evapotranspiration calculated from temperature-based potential evapotranspiration (PET), R is streamflow and ΔS is the change in soil water storage. Because groundwater fluxes are normally an order of magnitude smaller than other fluxes and due to the very low permeability of the organic soil, groundwater fluxes can be ignored (Skaggs et al., 1991). A previous study found deep, vertical seepage to aquifers in the Albemarle-Pamlico region of North Carolina to be less than 12 mm per year (Heath, 1975). Pocosins are topographical high points on the landscape and therefore also unlikely to receive groundwater input. The water balance equation provides a simple and useful means of establishing the hydrologic characteristics of a watershed at an appropriate temporal scale. (Dingman, 2002 and Vörösmarty, 1998). The usefulness of the water balance model for describing hydrologic characteristics of a watershed however will depend on accuracy of the input parameters, especially over shorter time periods (Xu and Singh, 1998).

The water balance equation provides a useful framework to test for watershed size by evaluating closure at multiple temporal scales. By assuming that closure always exists in the water balance, the watershed size determined by topography would be the same as the area actively contributing to runoff. Therefore, exploring the relationship between runoff measured at the watershed outlet and the topographically defined watershed size using the water balance could explain if a variable contributing area exists.

1.5. Research objectives and hypotheses

The main goal of this project was to investigate the key hydrologic components of a forested and ditched pocosin watershed in the ACP of North Carolina. The specific objective of this project is to develop and test a water balance framework for investigating runoff characteristics and the active watershed area contributing to runoff. More details of this objective and associated hypotheses are described below.

1.5.1. Water balance objective and hypothesis

Using a water balance approach, the aim is to calculate monthly and annual water balances with particular emphasis on deriving, mathematically, the area contributing to runoff.

Objective 1: To determine and analyze the water balance of the study site using the topographically delineated watershed area, observed precipitation, estimated evapotranspiration, observed runoff, and estimated change in soil water storage. Failure of the water balance to show closure (i.e. $P \neq R+ET+\Delta S$) is interpreted to mean that the active contributing area and the topographically delineated watershed area are not in agreement.

Hypothesis 1: There is closure (zero percent error) in the water balance

1.5.2. Watershed contributing area objective and hypothesis

Evaluating closure using observed and estimated variables in the water balance equation, and anecdotal evidence suggests that the area actively contributing to stream discharge at the watershed outlet (runoff) varies through time and may differ from the topographically defined watershed size (Objective 1).

Objective 2: To explore the concept of variable contributing area for a low relief coastal watershed. This was achieved by converting discharge measured at the outlet to runoff using a watershed area. The watershed area needed to produce runoff values that result in closure of the water balance equation represents the variable runoff contributing area. The calculation was done for different temporal periods:

monthly, seasonal, and annual and runoff contributing area values compared to the topographically defined watershed area.

Hypothesis 2: Under the assumption that water balance closure has been achieved, the active area contributing runoff within the watershed is constant and does not vary in response to the topographically delineated watershed boundary.

2. MATERIAL AND METHODS

2.1. Study area and site description

The study was carried out at the Yates Trail watershed located in Hofmann Forest, North Carolina, USA (34° 53'N, -77° 21'W). The Hofmann Forest was established by Dr. Julius "Doc" Hofmann in 1934 to support the research, demonstration and educational needs of the Forestry School. It has an extensive 70-year history of research activities and has undergone different management regimes from various groups. In 2008 management was transferred to the North Carolina Natural Resources Foundation, Inc. Today the Hofmann forest has a permanent staff and provides financial and research support to the Department of Forestry and Environmental Resources at North Carolina State University (NCSU).

The Yates Trail watershed forms the headwater of the south fork of the White Oak River, which flows eastwards to the Atlantic Ocean approximately 65km away (Figure 3). The difference in elevation between the highest and lowest points is approximately 2.5 meters, depending on whether the measurement was taken at ditch or ground surface levels.

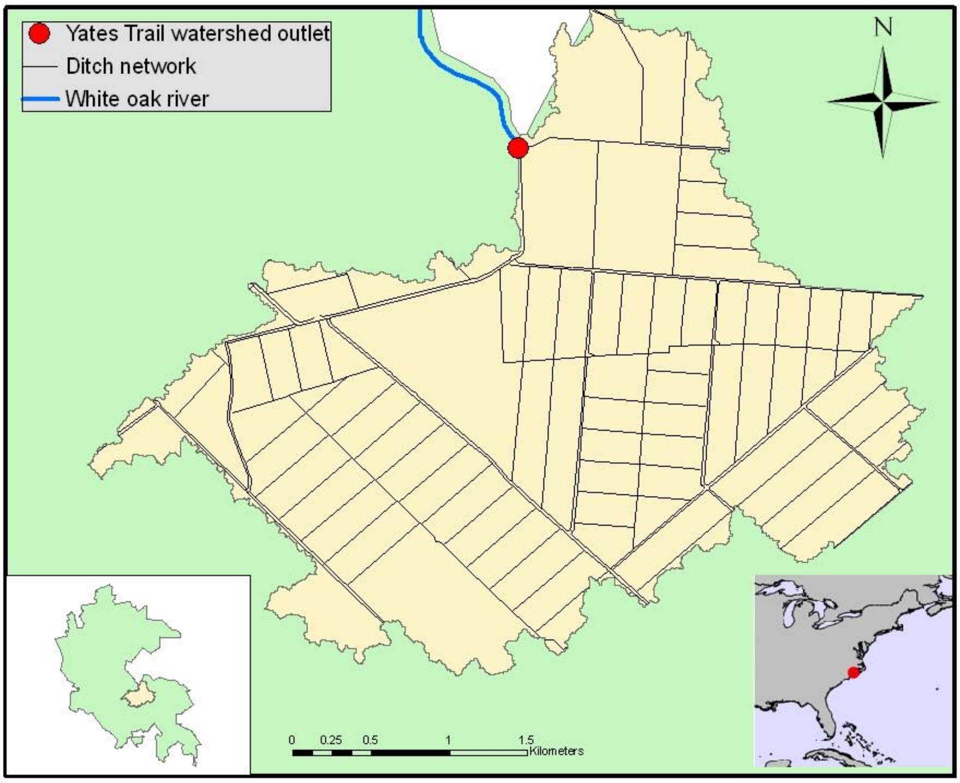


Figure 3: Yates Trail delineated watershed, ditch network within the watershed, and Hofmann Forest location

The Yates Trail watershed was chosen for this study among other watersheds in Hofmann forest because it has the most complete hydrologic dataset. Regular silvicultural and agricultural operations have been conducted in the watershed, including fertilization, timber harvesting and ditch maintenance (Table 1 and Figure 3). The impact of the management activities have not been specifically considered in this study. The forest now consists of a mosaic of loblolly (*Pinus taeda*), longleaf (*Pinus palustris*) and slash (*Pinus elliottii*) pine plantations at various rotation ages, native pocosin vegetation (Daniels et al., 1977) and linear patches of hardwood tree species. It is important to note that the ditch network (Figures 3 & 4) actually extends beyond the topographically delineated watershed boundary 1090 ha (see later).

Table 1: Forest management activities at the Yates Trail watershed.

Year	Final harvest (hectares)	Ditch (km)	Site preparation (hectares)
2006	63.3	-	-
2007	-	-	-
2008	58.9	-	-
2009	-	-	-
2010	-	4.3	-
2011	67.9	28.9	233.4
Sum	190.0	33.2	233.4

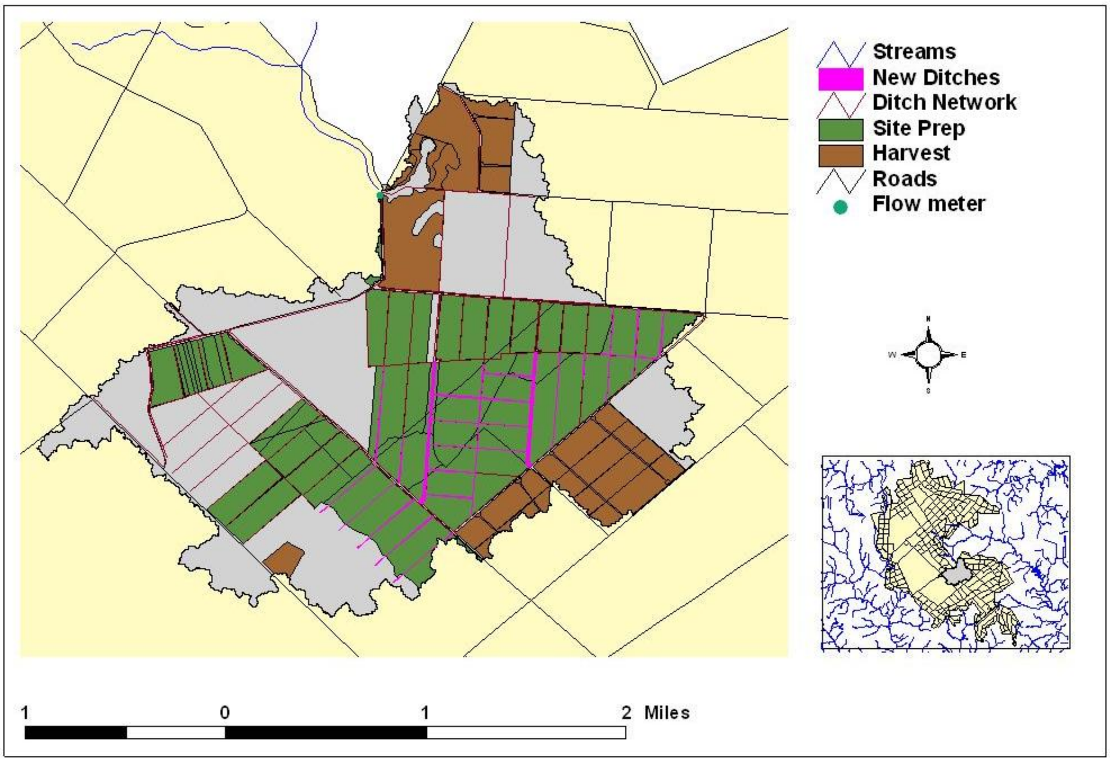


Figure 4: Yates Trail watershed (W_D) and distribution of landscape ditch network, infrastructure and instrument location.

Figure 5: Yates Trail watershed outlet and flow meter location (Source: Glenn Catts)

2.2. Watershed delineation

A watershed is a natural, topographically defined area draining to a single point in the landscape (Hornberger et al., 1998; Dingman, 2002). For this study, delineation of the watershed was carried out using spatial datasets of landscape variables and ArcGIS 9.3.1. A LIDAR bare earth, digital elevation model (DEM) with 20 cm resolution and 6.1 m pixel size was obtained from the North Carolina Floodplain-Mapping Program. The ArcHydro Toolbox in ArcGIS 9.3.1 was used to delineate the Yates Trail watershed (W_D) with the ditch network burned in to an elevation of minus infinity, forcing water to flow through specific pathways. The difficulties encountered during the watershed and ditch network delineation process are not uncommon in study sites with a low topographic relief: the LIDAR points did not always reach the actual ground surface because of water filled ditches or dense herbaceous vegetation (Zhang et al., 2008, Al-Muqdadi and Merkel, 2011), and the ditch network was complex. Culverts were considered as part of the ditch network, as water flowed from one roadside ditch to another. Combining field observations (e.g. made during ditch cleanout activities) with automatic delineation techniques is probably the most efficient and accurate means of delineating watersheds in flat, wet areas like the Hofmann Forest.

2.3. Hydrometric data

Hydrometereological data from Yates Trial were provided by Dr. Glenn Catts, professor and Hofmann Forest liaison to the NCSU College of Natural Resources. The data were collected as part of the ongoing long-term monitoring effort at the Hofmann Forest. The hydrologic data set from June 7, 2007 to April 6, 2011 were used for this study. Daily data were compiled into monthly, seasonal and annual datasets for further analysis.

A tipping bucket rain gauge (HOBO Data Logging Rain Gauge – RG3, Onset Inc., Bourne, MA) was used to record precipitation (*P*) at second intervals. Mean annual precipitation during the 45-month study period (July 2007 – March 2011) was 1,388mm. The precipitation measurements are assumed to be an areal average for the entire watershed and spatial variability in precipitation distribution was assumed to be negligible. An extreme precipitation event (hurricane) occurred in the fall of 2010 that was abnormal in size and duration. The storm event occurred during two months and outflow response was delayed into the second month. Therefore, daily precipitation and outflow data for the monthly dataset of the last 5 days of September 2010 were moved to October 2010 were excluded from analysis. Note that storm period analysis data were unmodified.

Stream discharge was measured using an automatic velocity type, flow meter (Stingray, Greyline, Massena, NY, U.S.A.) (Figure 6) placed in the outermost 2-foot diameter culvert at the watershed outlet (Figure 3).

a. b.

Figure 6: a) Greyline Stingray flow meter (Source: Greyline Instruments Inc.). b) Flow meter sensor installed flat against the bottom edge of the exit end of a corrugated aluminum culvert (Source: Dr. Glenn Catts).

The flow meter measured water level and velocity in ten-minute intervals from June 7, 2007 to April 6, 2011. The raw data were carefully inspected for outliers and gaps. These values were removed from the dataset when deemed reasonable to do so. Outliers were considered those values that were an order of magnitude larger than immediately adjacent values. Periodic gaps in data existed due to equipment failure. Discharge was calculated first in cubic feet per second (cfs) using water level height, cross-sectional area based on a circular culvert with diameter of 6 feet and velocity of the water flow. The cfs values were later converted to liters per second (L/s). The ten-minute interval discharge data were averaged to give daily values which were divided by the delineated watershed size (W_D), 1,090 hectares, to give “measured” runoff values (R_m) in mm per day.

A Remote Automatic Weather Station (RAWS) is located near the headquarters of Hofmann forest, approximately 10 km from the Yates Trail watershed site. The station measures daily air temperature, rainfall, humidity, wind speed, vapor pressure, and solar radiation. Daily weather data were used to calculate Hamon's potential evapotranspiration (PET_m, mm/month) (Lu et al., 2005) as

$$PET_m = 0.1651 \cdot D \cdot RHOSTAT \cdot KPEC \quad (2)$$

where D is day length (hr), the time from sunrise to sunset in multiples of 12 hours and calculated from the date, latitude, slope and aspect of the watershed; RHOSTAT is the saturated vapor density (g/m^2) at the daily mean temperature (TEMP) (ºC); and KPEC is the correction coefficient to adjust Hamon's PET values (a value of 1.2 was chosen, Federer and Lash, 1978). RHOSTAT was calculated as:

RHOSTAT = 216.7 · ESAT/(TEMP + 237.3) where; ESAT the saturated vapor pressure in mbars = 6.108 · exp[17.26939 · TEMP/(TEMP + 237.3)].

The Hamon temperature-based method of estimating evapotranspiration was chosen over the standard Penman-Monteith radiation method because it requires fewer data and better correlates to actual evapotranspiration (AET) at the watershed scale in the Southeastern United States (Lu et al., 2005). Actual evapotranspiration is the actual amount of water delivered to the atmosphere and lost by a vegetated surface (Dingman, 2002).

Due to the absence of RAWS weather station temperature data for January 2010, the ET value for this month was replaced with the MODIS (Moderate Resolution Imaging Spectroradiometer) value. MODIS ET data for the study site were obtained from the Global Subsetting and Visualization Tool provided by the NASA Oak Ridge National Laboratory (http://daac.ornl.gov/MODIS/MODIS-menu/MODIS_global_intro.html). The Hamon and MODIS ET values for the other months showed that MODIS ET was sufficient for the substitution (R^2= .93), especially during the wetter months (Figure 7). The high correlation between MODIS ET and Hamon's ET indicates a good relationship during the wetter months, but as months become drier MODIS ET drops below the level of Hamon's ET.

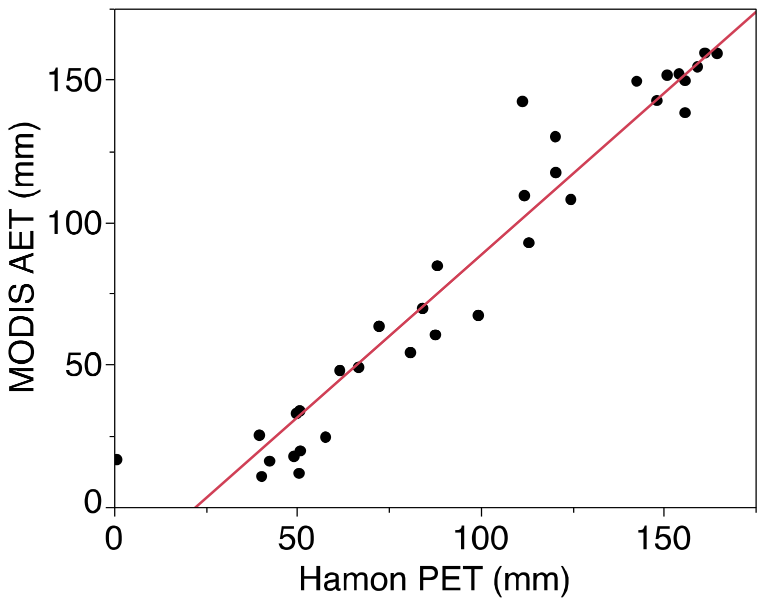


Figure 7: Monthly MODIS AET and Hamon's PET calculated from Hofmann Forest RAWs data.

Data from a continuously recording shallow groundwater monitoring well (Ecotone WM 1.0m water level monitor, RDS Inc., Navassa, NC) located near Sopp Hollow Road approximately 10 km (34.89435526, -77.45166575) from the Yates Trail watershed outlet, was used to calculate changes in Yates Trail watershed water storage. The well is more than 300 m away from the nearest ditch and therefore not influenced by the ditch network. The change in daily water table depth was multiplied by the drainable porosity to obtain the measured change in storage (ΔS_m). A drainable porosity value of 5% (cm cm^{-1}) was used based on a previous study conducted at similar ACP sites (Skaggs et al., 1991). Drainable porosity describes the volume (depth) of water that can drain from a given volume (depth) of soil by gravity.

2.4. Water balance closure

Because of incomplete daily hydrological data for June 2007 and April 2011 and problem with the data for September and October 2009 due to the hurricane (see earlier), there were 45 months with data for calculating the water balance. Closure of the water balance equation (Eqn. 1) was tested using measured and calculated water balance components (changes in soil water storage, runoff and ET) at monthly, seasonal and annual time scales. Calculated values for the water balance components were calculated as the residual term in the water balance equation. For example, changes in soil water storage (ΔS_c) were calculated as:

$$\Delta S_c = P - PET_m - R_m \tag{3}$$

Comparisons were made between measured and calculated water balance components using the closure error % (Mccarthy et al. 1991) and goodness-of-fit statistical criteria. Goodness-of-fit criteria used were coefficient of determination, R^2, and the Nash-Sutcliffe efficiency coefficient, R_{NS}. The Nash-Sutcliffe coefficient is a normalized value that compares the residual variance (noise) to the measured data variance (information). It indicates how well observed data fits to simulated data according to a 1:1 line; the closer to 1, the more accurate the model. Statistical analysis was conducted using JMP Pro 9 (SAS Institute Inc., Cary, NC, 1989-2010). The Nash-Sutcliffe efficiency coefficient [*Nash & Sutcliffe*, Journal of Hydrology 1970] for model bias is:

$$R_{NS}^2 = 1 - \left[\frac{\Sigma(V_m - V_P)^2}{\Sigma(V_m - V_{im})^2}\right] \tag{4}$$

where V_m is the measured monthly value, V_p is the predicted monthly value and V_{im} is the average measured monthly value over the 47-month study period. Closure errors in the water balance equation were calculated after McCarthy et al., (1991) as follows:

$$\%\ \text{Error of Closure} = \frac{\Delta S_c - \Delta S_m}{F} \times 100\% \tag{5}$$

where ΔS_c is the calculated change in soil water storage and where F is the system flux in millimeters as:

$$F = \frac{(P + R_m + PET_m + |\Delta S_m|)}{2} \tag{6}$$

The system flux is the system's total water flux, inflow or outflow) (McCarthy et al., 1991). A closure value of 0 % indicates that there is complete closure of the water balance and that there is no error in any of the measured water balance components. Values of closure error increasingly different from 0 % (positive or negative) indicate increasing error in one of the measured water balance components.
In this study, water balance equation closure error is likely to be due to an error in runoff (lack of correspondence between the topographically determined watershed area, the extent of the drainage ditches and source of runoff), and error in the change in water storage (use of ground water level data from a single well located in undrained conditions) or in both components.

Water balance parameter definitions are presented to guide the reader at the end of this document in Appendix 1.

2.5. Active contributing area

Preliminary analysis using the water balance equation provided evidence that measured runoff at Yates Trail was inconsistent with the topographically delineated watershed size. In order to investigate an error in the runoff component, the effect of using different watershed (runoff contributing source) area was examined. The runoff contributing area is defined as the area needed to produce closure in the water balance. Watershed areas (W_v) ranging from 800-1400 hectares in increments of 50 hectares were used to calculate monthly runoff (R_a). The monthly mean percent error closure values calculated using the different watershed area values were then plotted against watershed area. A second method of examining watershed area was to calculate the watershed active area index (W_{ai}):

$$W_{ai} = \frac{R_m}{(P - PET_m + \Delta S_m)} \tag{7}$$

The index seeks to identify when the watershed contributing area is expanding or contracting based on movement around 1. It is also possible to calculate the watershed contributing area (W_{ca}) using discharge (Q) as the amount of water available for runoff ($P - PET_m + \Delta S_m$):

$$W_{ca} = \frac{Q}{(P - PET_m + \Delta S_m)} \quad (\text{for } P - PET_m + \Delta S_m > 0; 0) \qquad (8)$$

Watershed contributing area is then plotted against the available water to show how the active contributing area varies. The condition (P-PETm-ΔSm > 0; 0) ensures only positive runoff values were considered. A median value of contributing area (Equation 9) was used to compare to the W_{ca} plot due to the wide distribution and frequent violation of the resulting values.

The final refinement for estimating watershed size utilized a regression method (W_r) by solving for precipitation in order to predict streamflow. The regressions were computed to examine temporal estimates of watershed size for the 47 month study period, by season and by year (Note using the storm subset, see below). Note that some years contained an incomplete 'water year' and so only complete years of data are included for analysis. Precipitation was the dependent or response variable and PET_m, ΔS_m and Q were predictor variables for the regression model. Predicted streamflow ($1/R_m$) from the regression model was used to estimate watershed area using the parameter estimate. All regression relationships were first evaluated for p-value significance (0.05), and all relationships revealed that the model is able to significantly improve the ability to predict streamflow (R_p).

The dataset for regression derived watershed size was broken into storms of 'no rain' periods of 4 days or more, resulting in 71 observations. The purpose for breaking the dataset into storm periods was to find the contributing area that provides water balance closure for each storm event and to eliminate the hurricane influence that spanned 2 months. Storm periods also remove the arbitrary overlap of months. More observations enhance the watershed size estimates by providing a better linear regression fit. In order to obtain the best statistical inference of watershed size, a population of watershed size estimates was derived by 'bootstrapping' the regression coefficients. The bootstrap approach allows for frequency distribution of the estimates calculated from the resamples, thus giving an estimate of the sampling distribution of the initial sample statistic.

Contributing area analysis parameter definitions are presented to guide the reader at the end of this document in Appendix 1.

3. RESULTS

3.1 Water balance

The evidence of watershed component imbalances is evident at the high water balance equation closure errors for some periods (Table 2). Among the different temporal water balance equation closures, seasonal closure, spring (an average of all spring months in the study period) was the poorest result. Based on the goodness-of-fit statistics, only streamflow (R_p) provided a reasonable result. ΔS_m fluxes are much smaller than other components and therefore do not fully explain failure to close the water balance. The high correlation and low Nash-Sutcliffe efficiency value for measured storage and calculated storage might be explained by a difference in the absolute values. Since measured storage is well stage times the chosen .05% drainable porosity, the absolute value difference is caused by the drainable porosity scaling parameter. The steep slope of the regression equation for measured versus calculated storage suggests that the value chosen for drainable porosity is too small, possibly an order of 6.5 times and true drainable porosity may actually be 30%. Groundwater monitoring well measurements taken outside the study watershed may also introduce uncertainty into ΔS calculations.

Table 2: Water balance components and closure errors for selected periods. Values are in mm, closure is %, and Q/P is a ratio of R_p and Precip.

Period	**Precip**	$\boldsymbol{R_m}$	$\boldsymbol{PET_m}$	$\boldsymbol{\Delta S_m}$	$\boldsymbol{\Delta S_c}$	$\boldsymbol{ET_r}$	$\boldsymbol{R_p}$	**Closure**	**Q/P**
2008	1173	137	1096	17	-60	1019	94	6.3	0.12
2009	1491	288	1110	7	93	1196	387	6.0	0.19
2010	1488	251	1130	-1	108	1239	357	7.6	0.17
2008-2010	1384	225	1112	7	47	1151	280	2.9	0.16
Spring	75	11	92	-2	-27	67	0	28.3	0.14
Summer	146	9	159	-2	-22	140	0	12.6	0.06
Fall	155	21	83	5	51	129	77	35.0	0.14
Winter	85	23	37	2	25	60	51	30.8	0.27
R^2					0.73	0.15	0.74		
R_{NS}^2					-45.60	0.42	0.38		
ΔS Regression	ΔS_c = -0.275 + 6.521* ΔS_m								
ET Regression	ET_r = 39.116 + 0.579 * ET_m								
R Regression	R_p = -20.395 + 3.633 * R_m								

3.2 Watershed characteristics

The rainfall, runoff and water stage at Yates Trail are seasonally cyclic with some variation among years. Rainfall displayed a seasonal pattern with elevated levels during the fall and winter months (Figure 8). The measured daily runoff contained a total of 37 days of null daily value data gaps.

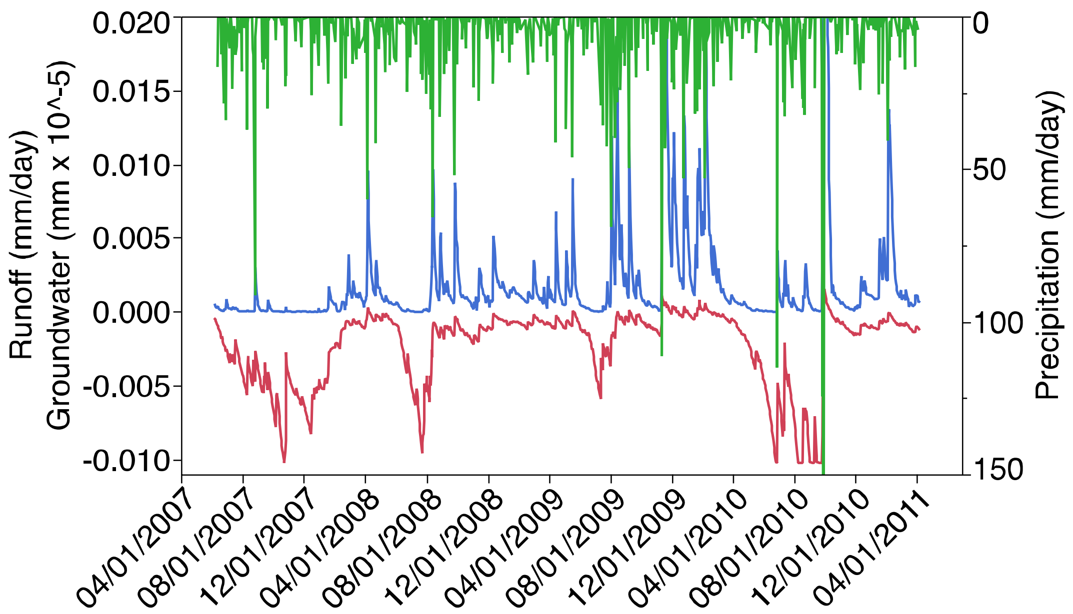
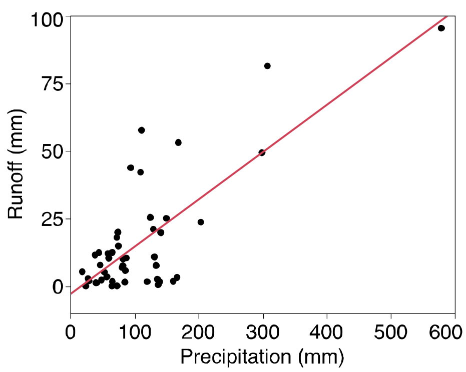


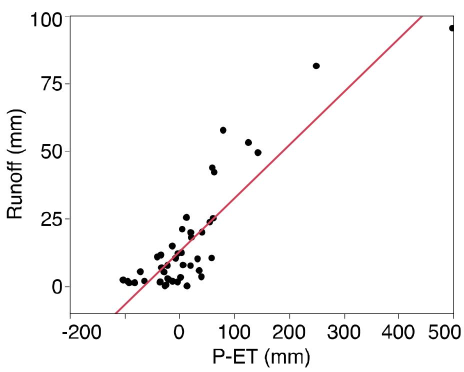
Figure 8: Daily hydrograph with groundwater stage (red) for the period June 6, 2007 to April 7, 2011. Runoff is shown in blue, and precipitation is shown in green.

In a nonparametric Spearman's ρ test, groundwater stage was found to be significantly related to runoff (Prob>| ρ | <.0001). Despite a near surface water table position at Yates Trail most of the year, during dry periods it can go deeper than 1m (Figure 8). The sum of yearly precipitation, as measured by the HOBO rain gauge, ranged from 1172mm in 2008 to 1490mm for 2010. Spatial variability of rainfall in this watershed is likely, but could not be captured without more extensive instrumentation or a spatially explicit remote sensing approach. Measured runoff from the Yates Trail watershed was highly variable by year and appeared to fluctuate by season, although seasons were not found to be significant in a Spearman's ρ test (Figure 8). The highest annual runoff occurred in 2009, and was twice as high as maximum outflow during 2008, a drier year.

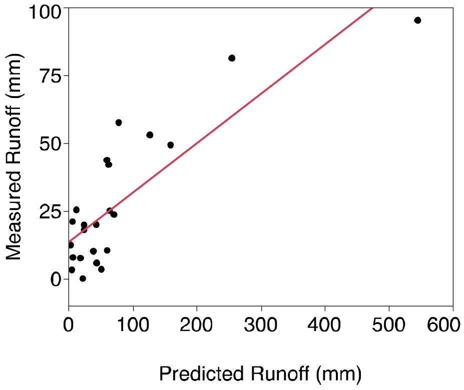
Streamflow and precipitation were correlated (R^2 = 0.58, R_m = -2.89 + 0.17*P) at the monthly time scale (according to monthly time steps) (Figure 9a). The steepness of the line indicates whether runoff is over or underestimated by precipitation at a given temporal scale (in this case monthly); this line suggests that runoff is overestimated. The runoff response results demonstrate that the variability in streamflow is not only a function of precipitation but also its temporal distributions (Figure 8). The relationship between runoff and precipitation is expected to be linear at longer time scales (season and year) and a good fit is an indicator of consistent response of runoff. Higher fall precipitation increases soil water storage and relatively low transpiration in winter increases soil water storage during this season as well. Seasonal runoff and seasonal precipitation were not found to be significantly correlated in any statistical tests, indicating runoff response is more variable at longer time scales. The runoff ratio at this scale is therefore sensitive to factors such as soil water storage, vegetation activity and storm intensity. As a refinement of the runoff and precipitation plot incorporates evapotranspiration; it provides a better fit (R^2 = 0.78, R_m = 12.88 + 0.19*[P-PET_m]) due to inclusion of the largest flux in the water balance equation (Figure 9b). The steep slope of the regression fitted line relative to 1 suggests that P-ET grossly overestimates runoff, especially in months when measured runoff is less than 25mm.



a.



b.



c.

Figure 9: a). R_m and precipitation, linearly fitted to evaluate consistency of runoff response to precipitation. b) R_m and precipitation minus PET_m. c) R_m and R_p. All figures computed using monthly time step data.

Finally, the remaining components of the water balance equation (Equation 1) were used to calculate residual runoff, and it is plotted against measured runoff (Figure 9c). The predicted runoff fitted well to measured runoff (R^2 = 0.74, R_m = 8.37 + 0.20$*R_p$), despite omitting values where closure was not achieved.). The outlier value (Figure 10c) was due to the hurricane event in the last days of September 2010. Runoff, runoff ratio (R/P) and change in soil water storage were plotted to observe the watershed signature (Figure 10). The dispersal or clustering of points shows the magnitude and frequency of responses of runoff to changes in watershed storage, which is soil water influenced by shallow groundwater. The dispersal of points about zero indicates storage fluctuates according to the frequency and magnitude of runoff (Figure 10. a. and c). A wider dispersal of points suggests that soil water storage is more a product of the relationship between runoff and precipitation (Figure 10. b. and d).

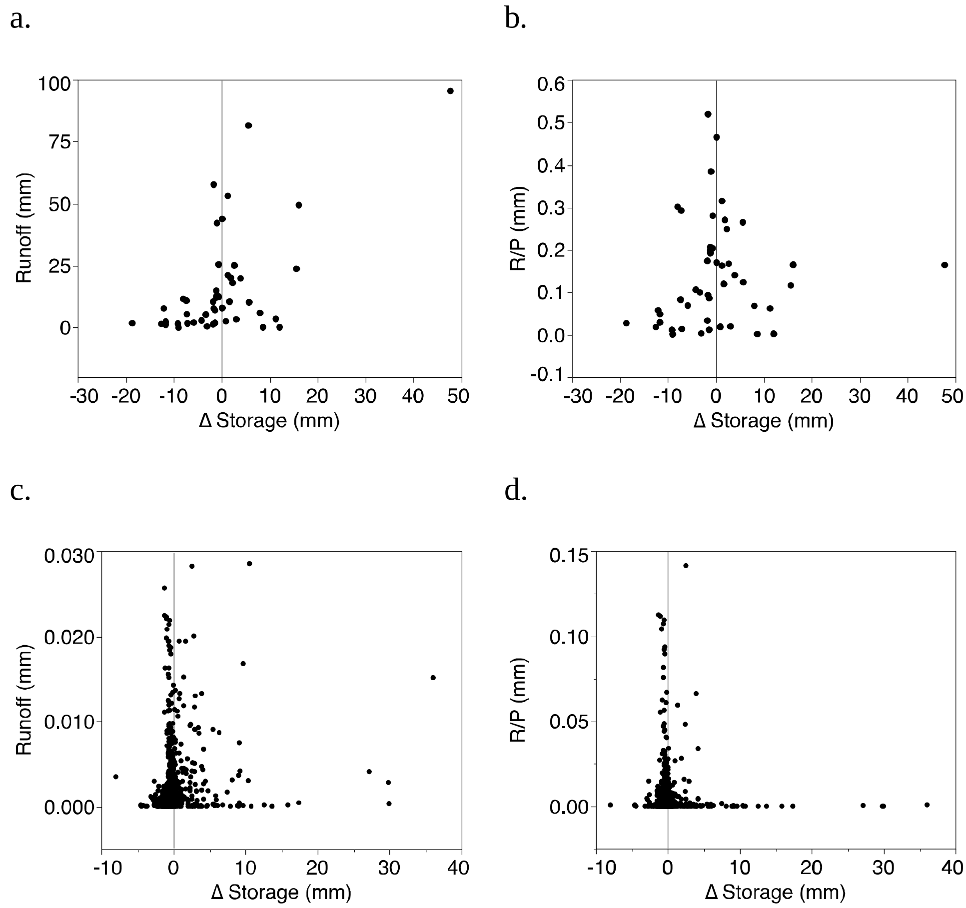


Figure 10: Monthly watershed signature, all values measured, monthly outlier storage value removed from October 2010, no outliers removed from daily data. a). Monthly ΔS and R_m. b). Monthly runoff/precipitation ratio and soil water storage. c). Daily ΔS and R_m d). Daily runoff/precipitation ratio and ΔS.

Monthly evapotranspiration residual (ET_r) was plotted against average monthly groundwater stage to assess if plant-available water is dependent on soil water capacity (Figure 11). The figure indicates that the correlation is poor and monthly average groundwater stage is unrelated to monthly ET_r. In a nonparametric Spearman's ρ test, no significance was found. Groundwater stage fluctuates down to a level of -800mm with seemingly no effect on ET. This might be explained due to conditions of the groundwater well outside the study watershed. The greatest drop in groundwater depth likely occurred seasonally in summer when plant water demand peaked and rainfall generally less frequent. This indicates that at the monthly time scale, plant water availability is not limited by depth of groundwater.

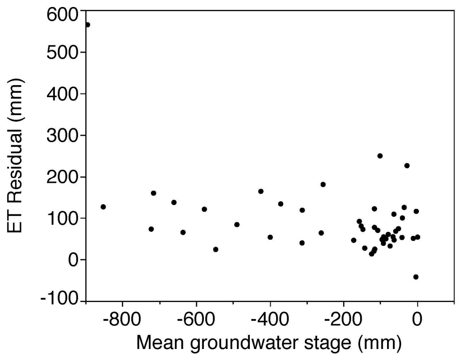


Figure 11: Mean monthly groundwater stage and monthly evapotranspiration residual (ET_r).

3.3 Active contributing area

The first approach (W_v) evaluated whether or not water balance closure could approximate size of the watershed (Figure 12a). A few distinct spikes are seen at 850, 1,000 and 1400 hectares, indicating a higher likelihood of the true value at those sizes. According to this method, when closure is 100%, the average, active watershed size is 600 hectares. The active area index (W_{ai}) suggests that the active area fluctuates regularly, in orders of magnitude both larger and smaller than W_d (Figure 12b). Seven months of the study period showed an expanding or contracting contributing area that was outside the normal distribution. The negative values of the index, although physically impossible, were included to illustrate scale and confirm occasions when PET_m exceeded precipitation. Average contributing area calculated from the monthly active area monthly time series was 330 ha. Solving for watershed contributing area (W_{ca}) indicated that most of the time water balance components are in agreement. The monthly streamflow plotted to monthly surplus provided a good fit, indicating that the relationships between other water balance components and streamflow are good (Figure 12c). This figure also shows average active area, the slope of the line and calculated from the regression equation (W_{ai}= 13.653 + 222.78 * X) results in a watershed size of 223 ha. As seen in the figure, some months have a larger watershed size than topographic size.

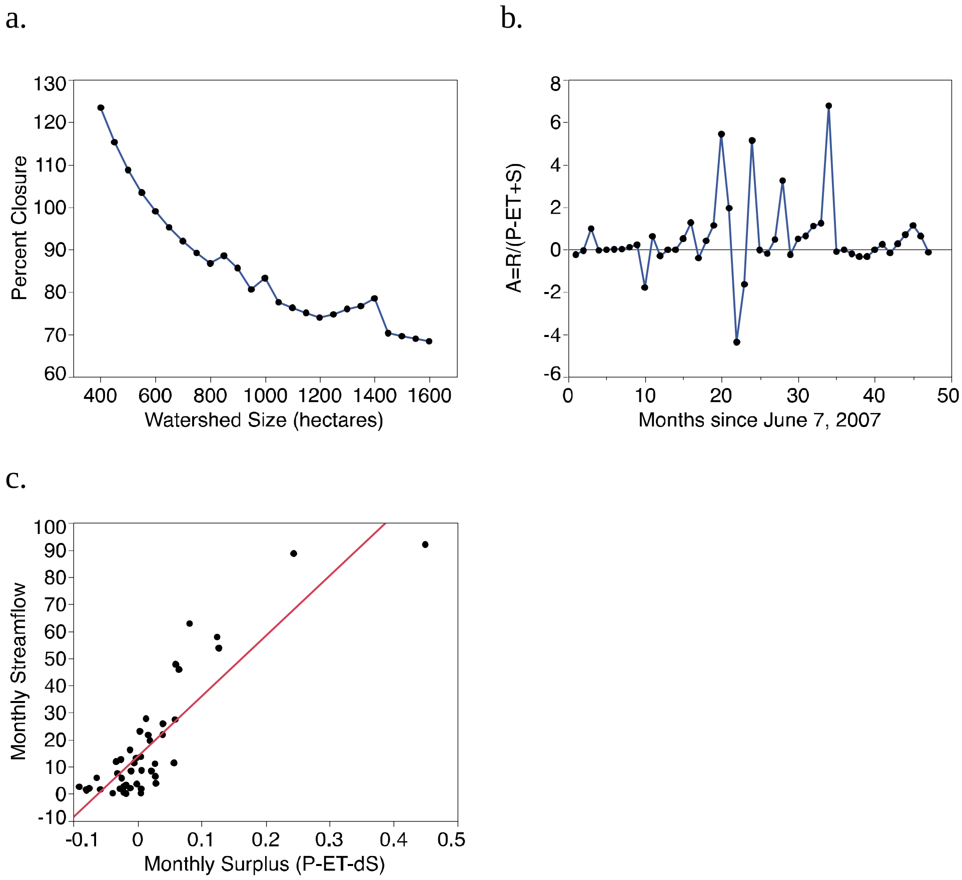


Figure 12: Watershed size plotted to water balance equation manipulations using monthly values (Equation 6). Drier conditions in autumn 2007 created near-zero runoff and caused area to plummet. a). Watershed size plotted to percent water balance equation closure. b). Active area monthly time series. c). Average active area.

The water balance equation regression approach (W_r) builds on the previous contributing area concepts to demonstrate the temporal variability in watershed size. Regression predicted watershed sizes are much smaller than the delineated watershed size (W_d) and contain a wide distribution of possible sizes (Table 3). A watershed size of 425 hectares was predicted for the 71 storm periods of the entire dataset and 506 hectares and 366 hectares for the lower and upper 95% confidence intervals, respectively (Regression model = -9.647 + 2.35E-13 * P). Periods 2008 and summer season resulted in unusually high confidence interval estimates, those parameter estimates were omitted (i.e. 2008). The bootstrap resample indicated smaller watershed sizes than W_d (Table 4). Most of the time (95%), watershed size can be expected around 450 hectares based on a change in precipitation.

Table 3. Water balance regression estimates and 95% confidence interval limits.

Time period	Active Area (ha)	Regression	Active Area Lower 95% (ha)	Active Area Upper 95% (ha)
2008	949.7	= 2.729 + 1.05E-13 * P	596.6	0
2009	338.5	= -6.807 + 2.95E-13 * P	305.4	380.4
2010	531.6	= -17.010 + 1.88E-13 * P	389.0	840.3
Fall	347.3	= -12.310 + 2.88E-13 * P	207.9	1054.9
Spring	813.0	= 2.311 + 1.23E-13 * P	616.5	1193.7
Summer	683.5	= -13.447 + 1.46E-13 * P	272.3	0
Winter	598.1	= -1.965 + 1.67E-13 * P	462.7	844.6

Table 4. Water balance bootstrap regression estimates and distribution statistics.

Range	Active Area (ha)
Quartile 100%	833.3
Quartile 75%	476.2
Median	416.7
Quartile 25%	400.0
Quartile 0%	333.3
Upper 95% (μ)	453.5
Lower 95% (μ)	425.5

4. DISCUSSION

4.1. Runoff response and storage relationships

Watershed behaviour was evaluated in terms of runoff response to precipitation because it is a useful and quantifiable method to describe the interactions between the functional watershed attributes. The runoff ratio can describe the variability in soil water storage because it is one of the main factors affecting runoff (Figure 10). A several day delay of runoff response to the hurricane event in autumn 2010 illustrates the storage capacity and low hydraulic conductivity at Yates Trail (Figure 8). If there is sufficient time between major storm events for groundwater stage to lower substantially, these events may produce little runoff as most precipitation is utilized for replenishing soil water lost as ET. However, higher initial groundwater stage position at or near the surface correlated to greater runoff amounts. The results for Yates Trail (Table 2) illustrate the range of potential annual streamflow (runoff coefficients of 0.12 and 0.18, respectively).

The surface detention and roughness of a pocosin site contributes to the low runoff amounts; this has also been seen in small-scale studies where site preparation activities were conducted (Amatya et al., 1997). On-going site preparation using mounded beds to elevate tree seedlings outside the zone of saturation artificially enhances the surface area for increased evaporation. Further, the extensive controlled drainage activities at Yates Trail have essentially increased the pore space available to soak up more rainfall by artificially lowering the water table and adjusting the pace at which water enters or leaves the soil profile. Runoff efficiency is also a function of catchment size. Generally, the larger the catchment size, the lower the efficiency. The large watershed area, high absorption capacity and low slope gradients at Yates Trail promote a slow runoff response.

Change in storage is a function of the changing water table depth and drainable porosity, fluctuating at variable temporal scales. A shorter scale, daily for instance, will not indicate a strong relationship between the antecedent moisture conditions (storage) and precipitation or runoff. However, at a longer temporal scale like monthly, the relationship may shift to be strongly linear because of a smaller overall change in storage (Figure 10. a. and c.). Analysis performed using storm events instead of months is also expected to indicate a linear relationship. The change in soil water storage is actually downplayed in this watershed because it only describes the measured change in groundwater stage; the reservoir capacity of the site is actually quite high as evidenced by observations of increased outflows from new ditch installation. Soil water at Yates Trail is depleted based on soil hydraulic conductivity and the predominantly organic soil substantially slows water movement despite relative high moisture levels throughout the year. Since soil water storage was calculated from groundwater stage (outside of the study site), the actual soil moisture is likely much different. One explanation of variation in soil moisture is the history of forest management activities at Yates Trail. Intensive tree growth rates could be driving groundwater stage lower on average and the ditch network ensures it stays relatively low. The estimate of drainable porosity in this study could have been substantially lower than the actual as indicated by low rates of water balance closure at all temporal scales (Table 2). A future water balance study at this site would benefit immensely from groundwater stage data collected within the study area. A direct comparison between the runoff ratio and change in soil water storage suggests that runoff is a function of storage with nonlinear characteristics (Figure 10).

Evapotranspiration residual plotted against the mean groundwater stage should show a close association because plant-available water is dependent on soil water capacity (Figure 11). However, there was a non-existent relationship between these variables, possibly indicating systematic errors in some other water balance components. This result is also contrary to the common belief that vegetation is able to exploit more water when the water table is elevated. Seasonal trends between water balance variables were expected because of interannual evapotranspiration fluctuations and periods of soil inundation or ponding. Evapotranspiration appears to have the greatest effect on groundwater stage by lowering it periodically during growing seasons (Figure 11). The variability in surface structure of a watershed this size likely introduces uncertainty into the evapotranspiration estimate due to the management-induced changes in area of surface soil evaporation and vegetation characteristics. Similar discrepancies in evapotranspiration estimation were noted by others and were attributed to the effects of harvesting, stand type, crown cover and canopy capacity (Amatya et al., 1997).

4.2. Active contributing area

According to W_v (closure-based watershed size estimate), water balance closure improves at smaller active watershed areas, but lessens as this area increases, suggesting that a much smaller area than W_d may be contributing to runoff at a given time (Figure 12a). The closures indicated by W_v are smaller than regression predicted watershed sizes (W_r), demonstrating that the average active watershed area as determined by runoff characteristics is smaller than W_d. Based on the W_v approach, a reasonable estimate of average active watershed area is about 600 hectares. However, using the entire dataset for the W_v approach may have smoothed out interannual variations because the uncertainty among all the water balance components was averaged. Precipitation and PET are the dominant hydrologic drivers, a gross underestimation of PET or spatial precipitation patterns could be causing systematic errors in balancing the other equation variables. The W_r approach is heavily weighted towards PET_m because it accounts for significantly more as an export variable when compared to the other components in the water balance (Q and ΔS_m).

A better estimate of contributing area should be expected from the W_r method because storm events are accounted for individually instead of smoothed over by monthly time steps. One possible explanation is the large difference between measured and predicted streamflow. The active area index (W_{ai}) demonstrated that the active contributing area has seasonal tendencies according to frequency and magnitude of storm events (Figure 12b). Despite the temporal variability in contributing area, the active area should conceivably be related to flow variability; as the average contributing area increases, the streamflow increases (Figure 12c) (Horwitz, 1978). This suggests that variability in runoff changes by season depending on contributing area expansion and contraction. The contributing drainage network is difficult to isolate, but a possible explanation is due to the ditch network drawing in water via a steeper hydraulic gradient. These new active parts of the watershed might drain the outer reaches of the watershed as well as new contributing areas. However, the fluxes generated by active areas may not always result in a catchment outlet response (Ambroise, 2004). Figure 12b seems to indicate that the entire active area is seldom engaged and would probably be less than the extent of available active areas. There are also temporal differentiations in active periods, variable due to prevailing conditions such as antecedent moisture conditions and likely the result of gaps in precipitation activity.

Forest management activities were unaccounted for in this study and management including vegetation removal could have increased soil water storage capacity at the surface. In addition, new ditching would have altered the hydraulic gradient, possibly forcing water to flow outside W_d. Ditch construction should be expected to steepen the hydraulic gradient and encourage higher streamflow volumes by drawing in water from a larger area and expanding the active contributing area.

Other characteristics at Yates Trail may be contributing to a lack of water balance closure and subsequent temporal variations in the active contributing area. The very low hydraulic gradient and muck-soil characteristics of the site promote a long soil water storage residence time that may not be captured in 4 years of data. Different parts of Yates trail likely exhibit higher soil moisture parts depending on the hydraulic gradient, thus could be expected to contribute runoff earlier or later from a storm. The W_d boundary of Yates trail is directly adjacent to several unmanaged areas without a ditch network. Lateral seepage could be occurring at the unbounded ditches that serve as part of W_d, drawing water away from the watershed. Water loss due to deep seepage was confirmed to be negligible by several deep drilling studies (Heath 1975; Riekerk et al., 1979).

5. CONCLUSIONS

Basic watershed attributes, a water balance approach and an active contributing area concept were presented in this study. The hydrometric dataset spanning 47 months provided a novel opportunity to evaluate all accounts of inflows and outflows of a watershed undergoing continuous forest management activities. The relatively slow runoff response, seasonality and complex storage capabilities are indicative of similar ACP sites (Amatya et al., 1996). The water balance components were generally in good agreement and closed with one another at longer time scales, decreasing for shorter periods. Active contributing area clearly varies temporally but on average is estimated to be approximately 600 hectares. Regression predicted watershed size was smaller than expected which could have been due to the difference between measured and predicted streamflow. The extent at which the active contributing area fluctuates may depend on the uncertainty of measured water balance components, especially change in soil water storage.

Forest management activities including large area harvest, ditch installation and road building were not accounted for at all in this study and could have clouded the results. Future studies, using a similar dataset, could examine individual storm characteristics and known dates of forest management activities to isolate human influences as they relate to stream outflow characteristics. Newly available spatially explicit precipitation data (e.g. radar) should be obtained to evaluate the spatial distribution of precipitation and how it affects runoff dynamics. The theoretical framework of this study employing the water balance can be leveraged in future studies by honing in on relevant data parameters. The field of watershed hydrology has sophisticated tools available to test individual water balance parameters for their role in predicting watershed size, but can only be realized with a deliberately collected hydrometric dataset.

As of this thesis publication, long-term ownership and management of Hofmann Forest is uncertain. Regardless of future land management objectives at the site, continued monitoring efforts are essential to support the wide-ranging research interests of faculty and students at North Carolina State University.

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7. APPENDIX 1

AET: Actual evapotranspiration, the actual amount of water delivered to the atmosphere and lost by a vegetated surface (Dingman, 2002).

$\boldsymbol{ET_r}$: Residual evapotranspiration as the result of the water balance equation.

KPEC: the correction coefficient to adjust Hamon's PET values.

$\boldsymbol{PET_m}$: Measured potential evapotranspiration using the Hamon method, the amount of evapotranspiration that would occur if sufficient water were available.

$\boldsymbol{R_a}$: Calculated monthly runoff as a result of calculations for W_v.

$\boldsymbol{R_m}$: Measured runoff values in mm per day, based upon the delineated, topographically defined watershed area.

$\boldsymbol{R_p}$: Predicted runoff as result of calculations for water balance regressions and determination of W_r.

$\Delta\boldsymbol{S_m}$: Measured soil water storage from groundwater stage data.

$\Delta\boldsymbol{S_c}$: Change in soil water storage, calculated as the result of $\Delta S_c = P - PET_m - R_m$.

$\boldsymbol{V_m, V_p, V_{im}}$: Measured monthly value, predicted monthly value and average measured monthly values over the study period used to calculate the Nash-Sutcliffe efficiency coefficient.

$\boldsymbol{W_{ai}}$: Watershed active area index, see equation 7.

$\boldsymbol{W_{ca}}$: Watershed contributing area, see equation 8.

$\boldsymbol{W_D}$: The topographically defined watershed area as determined by conventional watershed delineation.

$\boldsymbol{W_r}$: Regression method for predicting watershed size from P and Q.

$\boldsymbol{W_v}$: Watershed area adjusted by size to produce closure in the water balance.